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PLANETARY BOUNDARY LAYER PARAMETERIZATION FOR TROPICAL CYCLONES--ETC(U)
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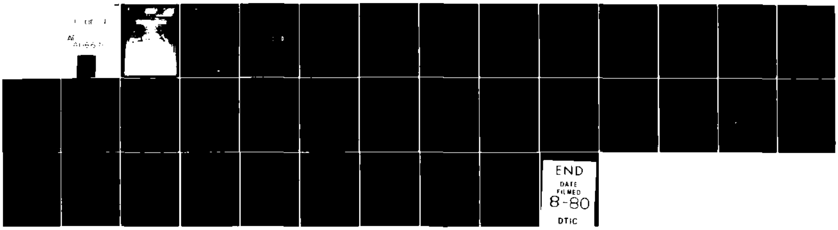

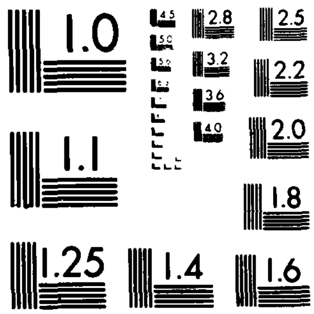

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20. ABSTRACT (Continue on reverse side if necessary and identify by block number)

A planetary boundary layer (PBL) parameterization base on the generalized similarity theory (GST) was developed for tropical cyclone models. This parameterization, with only one layer, is necessary in modeling tropical cyclones for computational speed. The momentum, sensible heat and moisture fluxes are mutually dependent in this parameterization through non-dimensional gradient equations. The internal structure of the PBL is determined implicitly through universal functions.

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20. Abstract (Continued)

In comparison with a complex, one-dimensional, multi-layer PBL model, the GST parameterization yields accurate moisture fluxes, but slightly overestimates the momentum flux and underestimates the sensible heat flux. The GST parameterization produces very realistic dynamics, energetics and thermal structure in an axisymmetric tropical cyclone model. This GST parameterization is judged superior to drag coefficient parameterization and is a good alternative to the more expensive, multi-layer parameterization.

CONTENTS

PLANETARY BOUNDARY LAYER PARAMETERIZATION FOR TROPICAL CYCLONES BASED ON GENERALIZED SIMILARITY THEORY

I. INTRODUCTION

The planetary boundary layer (PBL) plays a critical role in the evolution of tropical cyclones because the air-sea energy and momentum exchanges occur through the PBL. Parameterization of the PBL is thus a very important aspect of tropical cyclone modeling. Complex parameterization is not always computationally feasible for three-dimensional or operational models. Computationally efficient yet accurate PBL parameterization is imperative for these models. The purpose of this paper is to test such a parameterization based on the generalized similarity theory (GST) in the parameterization of the PBL in tropical cyclone models.

The most widely used PBL model is the single-layer parameterization where the PBL is represented by one model layer and the surface fluxes are calculated using bulk aerodynamics formulas. This method is simple and computationally economical but has a major drawback in that the internal structure of the PBL cannot be resolved. As a consequence, the single layer PBL model using bulk aerodynamic formulas can only crudely estimate various surface fluxes.

An alternative approach where the PBL is modeled by several layers has been adopted in recent years by Pielke (1974), Kurihara and Tuleya (1974), Anthes and Chang (1978) and Anthes and Warner (1978). By explicitly resolving the PBL structure, these models often yield more realistic estimates of the various fluxes, but the computational cost is high.

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Recent observational studies have facilitated the testing of various PBL similarity theories. The rather idealized assumptions characteristic of early theories have been relaxed and improved, and more realistic and general theories have been derived (Arya 1977; Yamada, 1976). These developments have led to a PBL formulation that appears suitable for various PBL conditions.

In the next section, a brief review on the generalized similarity theory will be given. In Section 3, various surface fluxes in typical tropical cyclone conditions obtained from the GST will be compared with those computed by a time dependent one dimensional PBL model (Busch, et al., 1976; hereafter referred as BCA). The generalized similarity theory is then applied to an axisymmetric tropical cyclone model. The results will be compared in Section 4 with those obtained using constant drag coefficients.

II. Generalized Similarity Theory

Similarity theory states that the profile of any properly scaled variable in the PBL such as wind, temperature, or water vapor, can be described by a universal function. Although there has been much debate concerning the proper scales for the various types of PBL's, we will use the similarity theory that appears most applicable and general for tropical cyclones (Ayra, 1977, 1978; Yamada, 1976).

The mathematical representation of the PBL flow can be obtained by assuming that there exists a layer where both the equations describing the surface layer and the equations describing the interior flow in the mixed layer are both valid. In the surface layer, the PBL variables are assumed to follow a logarithmic-linear relationship with height

$$\frac{|\underset{\sim}{v}|}{u_*} = \frac{\hat{t}}{k}\left[\ln\left(\frac{z}{z_o}\right) - \Psi_m\left(\frac{z}{Z}\right)\right] \tag{1a}$$

$$\frac{\theta-\theta_o}{\theta_*} = \frac{P_r}{k}\left[\ln\left(\frac{z}{z_o}\right) - \Psi_h\left(\frac{z}{L}\right)\right] \tag{1b}$$

$$\frac{q-q_o}{q_*} = \frac{P_r}{k}\left[\ln\left(\frac{z}{z_o}\right) - \Psi_q\left(\frac{z}{L}\right)\right] \tag{1c}$$

where $\hat{t}$ is the unit vector parallel to the surface wind, P_r is the turbulent Prandtl number, Ψ_m, Ψ_h and Ψ_q are stability functions, z_o is the roughness length, and θ_o and q_o are potential temperature and water vapor content at z_o.

For the interior flow in the mixed layer we can write the equations in the form of the resistance law

$$\frac{\underset{\sim}{V}-\underset{\sim}{V}_m}{u_*} = \hat{t} \cdot F_u(\frac{z}{h}, \frac{h}{L}) + \hat{n} \cdot F_v(\frac{z}{h}, \frac{h}{L}) \text{ sign } f \tag{2a}$$

$$\frac{\theta-\theta_m}{\theta_*} = F_\theta (\frac{z}{h}, \frac{h}{L}) \tag{2b}$$

$$\frac{q-q_m}{q_*} = F_q (\frac{z}{h}, \frac{h}{L}) \tag{2c}$$

where V_m, θ_m and q_m are the mean velocity, potential temperatures and water vapor content in the PBL, $\hat{n}$ is a unit vector normal to $\hat{t}$, and h is the height of the PBL.

By matching (1) and (2), we obtain a set of equations that is simultaneously valid for surface layer and the mixed layer,

$$\frac{ku_m}{u_*} = \ln (\frac{h}{z_o}) - A(\frac{h}{L}) \tag{3a}$$

$$\frac{kv_m}{u_*} = - B(\frac{h}{L}) \text{ sign } f \tag{3b}$$

$$\frac{k(\theta-\theta)_m}{P_r\theta_*} = \ln (\frac{h}{z_o}) - C(\frac{h}{L}) \tag{3c}$$

$$\frac{k(q-q_m)}{P_r q_*} = \ln (\frac{h}{z_o}) - D(\frac{h}{L}) \tag{3d}$$

where u_m and v_m are the components of the mean velocity in the PBL in t and n directions, respectively. Universal functions A, B, C and D determined empirically. It is generally assumed that $D \approx C$.

Using equation set (3), and the Wagara observational data, Yamada (1976) obtained functions A, B and C (eqs. (13) through (18) in his paper) which are adopted here because of the simple functional form and accuracy.

It should be noted that $\underset{\sim}{V}_m$ in Yamada's analysis is the vertically averaged geostrophic wind in the PBL. Here $\underset{\sim}{V}_m$ is the vertically averaged mean wind. Thus the baroclinic effects are minimized (Arya, 1978). In addition, when actually applying Eq. (3) in a dynamic model, the sum of the squares and the quotients of Eqs. (3a) and (3b) are used to evaluate the magnitude and direction of the surface wind stress.

III. Generalized Similarity Theory and Multi-Layer Approaches in a One Dimensional Tropical Cyclone Model.

As stated in the introduction, the Parameterization of the PBL based on the GST for a tropical cyclone model is very appealing because it requires only one layer to represent the PBL. Thus it is important to test the universal functions described by Eq. (3) with the observations of tropical cyclones. Unfortunately, detailed observations of tropical cyclone PBL's are not yet available, so we resort to use a one-dimensional, multi-layer PBL model for comparison. Table 1 lists various models used in this study.

Table 1. Model Used for Comparison

Models	Type	Dimensions	PBL parameterization
GST	PBL	1	Single-layer, generalized similarity theory
BCA	PBL	1	Multi-layer, Busch et al. (1976)
S	Tropical Cyclone	2	Single-layer, generalized similarity
D1	Tropical Cyclone	2	Single-layer, $C_D = 0.0015$ $C_E = C_H = 0.003$
D2	Tropical Cyclone	2	Single-layer, $C_D = 0.003$ $C_E = C_H = 0.003$

The BCA 1-D PBL model can reproduce observations accurately and it has been used successfully in parameterizing the PBL in numerical models (Anthes and Chang, 1978; Anthes et al., 1978). The model has a surface layer in which the logarithmic-linear relationships for wind and temperature (Businger et al., 1971) are adopted. It should be noted that the same logarithmic-linear relationship is used by Yamada (1976) to generate the universal functions. Above the surface layer, a time dependent mixing length is used. The mixing length is assumed to continuously approach its asymptotic value, which depends on the depth and stability of the PBL. The e-folding time of the mixing length is proportional to the size of the energy-containing eddies in the PBL and inversely proportional to the turbulent kinetic energy.

The current version of the BCA model has 20 layers using a stretched grid from the surface to 3 km. The lowest layer defined as the surface layer, has a thickness of 25 m. To properly simulate tropical cyclone conditon, a rather strong pressure gradient force is specified so that the geostrophic wind speed is strongest at the surface and decreases with height to 90% of its surface strength at 3 km. The initial thermal state which has been taken from the model hurricane of Anthes and Chang (1978) shows a well mixed PBL below a capping inversion at 600 m with a sharp increase of potential temperature and decrease of specific humidity above the inversion. The initial wind is equal to the geostrophic wind prescribed for each run.

The BCA model was integrated 12 hours for 24 cases with the geostrophic wind varying from 12 to 35 m s^{-1} and the sea-surface temperature (SST) varying from 24.4 to 29.4^{o}C. These variations produce different stability conditions and large rates of changes in the initial hours. Diagnoses using Eq. (3) (or GSA model) were given at model times of 1, 2, 3, 4, 6, 8, 10, and 12 h. The vertically integrated u, v, θ, and q in the BCA model are the mean PBL properties needed for the diagnoses. Figure 1 shows the comparison of the surface wind stresses between the diagnoses from GST model and the computation of BCA model. The GST model produced nearly the same results (points on the diagonal line in Figure 1) as the BCA model for small u_*^2 value and for the unstable conditions. The stress diagnosed by the GST are higher than the BCA model under stable conditions. The overestimate by the GST model lessens as instability increases.

The larger estimation of the surface stresses for stable conditions of the GST model is due to the greater wind shear in the lower PBL which is resolvable in the BCA model. This strong wind shear is not properly represented by the universal functions in the single layer approach. When such a strongly sheared layer, which usually accompanies a low-level inversion, is present, the PBL mean wind may not be a proper velocity scale for the PBL. Fortunately, the formation of such a low-level inversion is not likely in tropical cyclones, the GST model therefore seems to estimate the momentum fluxes in typical tropical cyclones quite well.

Figure 2 and Figure 3 show the comparisons of the upward and downward sensible heat fluxes, respectively. For unstable conditions,

the GST model underestimates the upward heat flux because of the formation of a strong superadiabatic lapse rate region near the surface in the BCA model. For stable conditions the GST model overestimates the downward heat flux. This is related to the formation of low-level inversion in the BCA model which effectively cuts off the downward transport of heat. These characteristics of the GST model are found in all single-layer parameterizations.

Fig. 4 shows the comparison of the moisture flux $\overline{w'q'}$. The GST model gives a higher $\overline{w'q'}$ for the unstable condition, and a very similar $\overline{w'q'}$ for near neutral conditions, and a scattering in Fig. 4 is similar to that in Fig. 1, the more symmetric distribution about the diagonal in Fig. 4 is presumably related to the larger value of $\frac{\partial q}{\partial z}$ in the lower part of the BCA model.

These deviations of the GST model from the BCA model are much less than the model using aerodynamic drag coefficients. A supplementary computation (not shown) indicates various fluxes depend mostly on the wind speed only. The deviations of the GST model from the BCA model is partially inherited from the scattering of the original observation data from which the universal functions are deduced (Yamada, 1976). The comparison, however, does indicate a distinct behavior of the GST model for different stability. It should be pointed out the the GST has been applied to highly non-stationary conditions in this analysis. We note that as the PBL approaches a quasi-steady state, the GST model diagnoses becomes comparable to the BCA model results.

The experiments with the one-dimensional model indicates that the GST model agrees very well with the Busch model in momentum and moisture flux predictions. The GST model does not predict accurate fluxes when there are strong low-level gradients of potential temperature. The heat flux, however, is fortuitously not critically important in the development and maintenance of tropical cyclones as pointed out by Rosenthal (1971) and Anthes and Chang (1978). The single layer approach with GST model in parameterizing the PBL in a tropical cyclone is thus adequate.

IV. Incorporation with an Axisymmetric Tropical Cyclone Model

a. Model Review

The axisymmetric tropical cyclone model used for our investigation is similar to the one described in Anthes and Chang (1978) except that (1) the finest resolution at small radii is 30 km, and (2) there are six vertical layers. The vertical σ-layers are bounded at levels σ = 0.0, 0.2, 0.3, 0.6, 0.8, 0.93, and 1.

Included in one earlier version of the model was a time dependent equation for the PBL height (Deardorff, 1972). Supplementary experiments showed that the equation predicts infinite growth of PBL near the center of the cyclone unless large horizontal diffusion is applied. We, therefore, hold the height of PBL constant between σ = 0.93 and 1. The PBL has a depth of 650 = 700 m, a typical observed PBL depth near topical disturbances (Moss and Merceret, 1976; Nitta, 1974). Charnock's equation (Delsol et al., 1970) is used to compute the roughness length. The initial conditions consist of a weak vortex, which is in gradient balance, with maximum tangential velocity of 17 m s^{-1},embedded in a tropical atmosphere (Jordan, 1958). The lateral boundary conditions are Dirichlet for the thermodynamic variables and zero-vorticity for the momentum.

For comparison purposes models were also integrated using constant drag coefficients with C_D = 0.0015, C_H = 0.003 (model D1) and with C_D = 0.003, C_H = C_E 0.003 (model D2).

b. Development into Tropical Cyclones

The development from the initial vortex into a tropical cyclone is depicted by the maximum surface wind (Fig. 5) and miminum pressure (Fig. 6). After the initial dissipation due to friction, Model S rapidly develops into an intense cyclone and reaches a quasi-steady state at about 40 hr. The maximum surface wind is about 57 m s^{-1} and the minimum surface pressure is approximately 958 mb.

Models D1 and D2 also intensify after the initial dissipation stage. These two models deviate considerably from each other and from Model S after 20 h. Model D1 reaches a final intensity of 35 m s^{-1} and 986 mb, whereas Model D2 reaches a final intensity of 43 m s^{-1} and 980 mb.

c. Cyclone Structure at Quasi-Steady State

The tangential velocity field at 48 h for Model S shows a strong and concentrated cyclonic circulation near the center (Fig. 7). The maximum velocity is at r = 30 km in the lowest level and the 40 m s^{-1} contour extends upward to $\sigma = 0.6$. The tangential circulation diminishes upward due to the baroclinic effect of the warm core and anticyclonic circulation occurs at the $\sigma = 0.2$ level outside of 300 km. The inflow is confined to the PBL with maximum radial velocity of 35 m s^{-1} just outside the region of the maximum tangential velocity (Fig. 8). Strong outflow with velocity more than 20 m s^{-1} occurs at the $\sigma = 0.2$ level. Except for the shallow inflow and outflow layers, the radial velocity is small in mid-troposphere.

Figure 9 shows the relative humidity (RH) at 48 h of Model S. The RH field features (1) very moist PBL and outflow canopy, (2) relative moist convective eyewall region, (3) a dry eye region due to the strong sinking motion, and (4) a dry mid-troposphere outside the eyewall due to the general subsidence.

The temperature field of the quasi-steady tropical cyclone shows a warm core with an anomaly of more than + 12^{o}C. The + 8^{o}C anomaly extends to r = 300 km in the outflow layer. The + 2^{0} C anomaly contour between σ = 0.6 and σ = 0.9 outlines the subsidence. The cooler PBL near the center is due to the adiabatic expansion (and therefore cooling) in the inflow.

d. Air-Sea Energy Transfer

We will use budget calculations of kinetic energy, moisture, and heat to compare model results and reveal the importance of various physical processes. As shown in Anthes and Chang (1978), the budgets of kinetic energy (K), moisture (Q) and enthalpy (H) for a tropical cyclone model can be expressed schematically by:

$$\frac{\partial K}{\partial t} \approx K_b + C + D_v + D_L + D_H$$

$$\frac{\partial Q}{\partial t} \approx Q_b + E + P + Q_H$$

$$\frac{\partial H}{\partial t} \approx H_b + A + Q_p + H_s + H_H$$

where the terms with subscript b's denote the lateral boundary fluxes, and terms with subscript H's denote the effects due to horizontal diffusion. In addition, D_V, and D_C represent the rates of kinetic energy dissipation due to surface friction and cumulus friction respectively. E and P denote evaporation and precipitation rates respectively. Q_P represents latent heat release, and H_s the surface sensible heat transfer. A denotes the adiabatic heating rate due to vertical motion; C denotes the kinetic energy conversion rate due to cross-isobaric flow.

Within the model domain of σ = 0 to 1 and r = 0 to 300 km, the most direct contribution of the PBL processes are the air-sea sensible and latent heat exchanges. Figure 11 and 12 show the time series of H_s and E. During the early stages, the sensible and latent heat transfers of D1 are equal or greater than those of S. The sensible and latent heat exchanges in S, however, grow steadily after 10 hr and finally reach 15 x 10^{12} W and 3 cm d^{-1} respectively, whereas in D1, the sensible and latent heat exchanges only grow to maxima of 7.5 x 10^{12} W and 1.8 cm d^{-1} after 30 h. After the maxima are reached, they decrease with time because of stabilization and saturation of the PBL.

Apparently, the sensible and latent heat exchanges in S are maintained at high levels primarily due to their nonlinear dependence on wind speed and stability. As wind speed in S approaches its maximum after 30 h, both H_s and E are enhanced to a greater extent than in D1. In addition, the radial circulation in S is also increased so that a slightly unstable and dry PBL is maintained by the strong inflow.

The increased radial circulation is illustrated by the conversion rate (C) from the available potential energy to kinetic energy shown in Fig. 13. The conversion rate is approximately proportional to the combined strength of the radial circulation and the warm core. It is apparent that the conversion rate in S grows rapidly between 10 to 20 h, which coincides with the increased evaporation. The proportionally faster increase of C than H_s and E in Exp. S shows that the increased heat transfer creates a stronger inflow, which, in turn, maintains a stronger heat transfer and a stronger inward transport of the dry, cooler ambient air.

The two different PBL parameterizations also gives different dissipation rates (Fig. 14). Note that D_V in S is smaller before 15 h than in D1, and it eventually asymptotes to three times stronger. Again, this is due to the nonlinear dependence of the surface stress on wind speed.

One can argue that the nonlinear dependence of the air-sea energy exchange could be achieved by employing wind-speed-dependent drag coefficients. However, the parameterization based on similarity theory is superior. The heat and momentum transfers depend not only on wind speed but also on PBL stability, and it is through PBL stability that various fluxes are correlated. The surface roughness appears in the formulation as a length scale and thus the parameterization captures an important feedback from the sea state. The turning of wind in low levels, neglected in the parameterization with drag co-

efficient, it also included implicitly. Therefore, the surface stress backs from the direction of the PBL mean wind, as expected.

Overall, the generalized similarity theory appears to be applicable in parameterization of the PBL in the tropical cyclone models. Such parameterization contains several important characteristics of the PBL such as the mutual dependence of various fluxes and the backing of the low-level wind, it also produces very realistic structures and energetics of tropical cyclones.

V. Summary

A PBL parameterization based on the generalized similarity theory has been tested for tropical cyclone models. We conclude that a single layer approach in parameterizing the PBL is viable although it cannot resolve the detailed structure of the PBL.

This parameterization was tested with the 1D multi-level PBL model of Busch et al. (1976) under various stabilities and highly non-stationary conditions. For stationary conditions, the computed water vapor flux, agrees well with the Busch model. The momentum flux is overestimated and the sensible heat flux is underestimated especially when there is a low-level inversion.

An axisymmetric tropical cyclone model which incorporated this single-layer-parameterization produced very realistic dynamic and thermodynamic structures. In comparison with otherwise identical tropical cyclone models that employ constant drag coefficients, the superiority of stress-dependent drag coefficients in the development of tropical cyclones was demonstrated.

Acknowledgments

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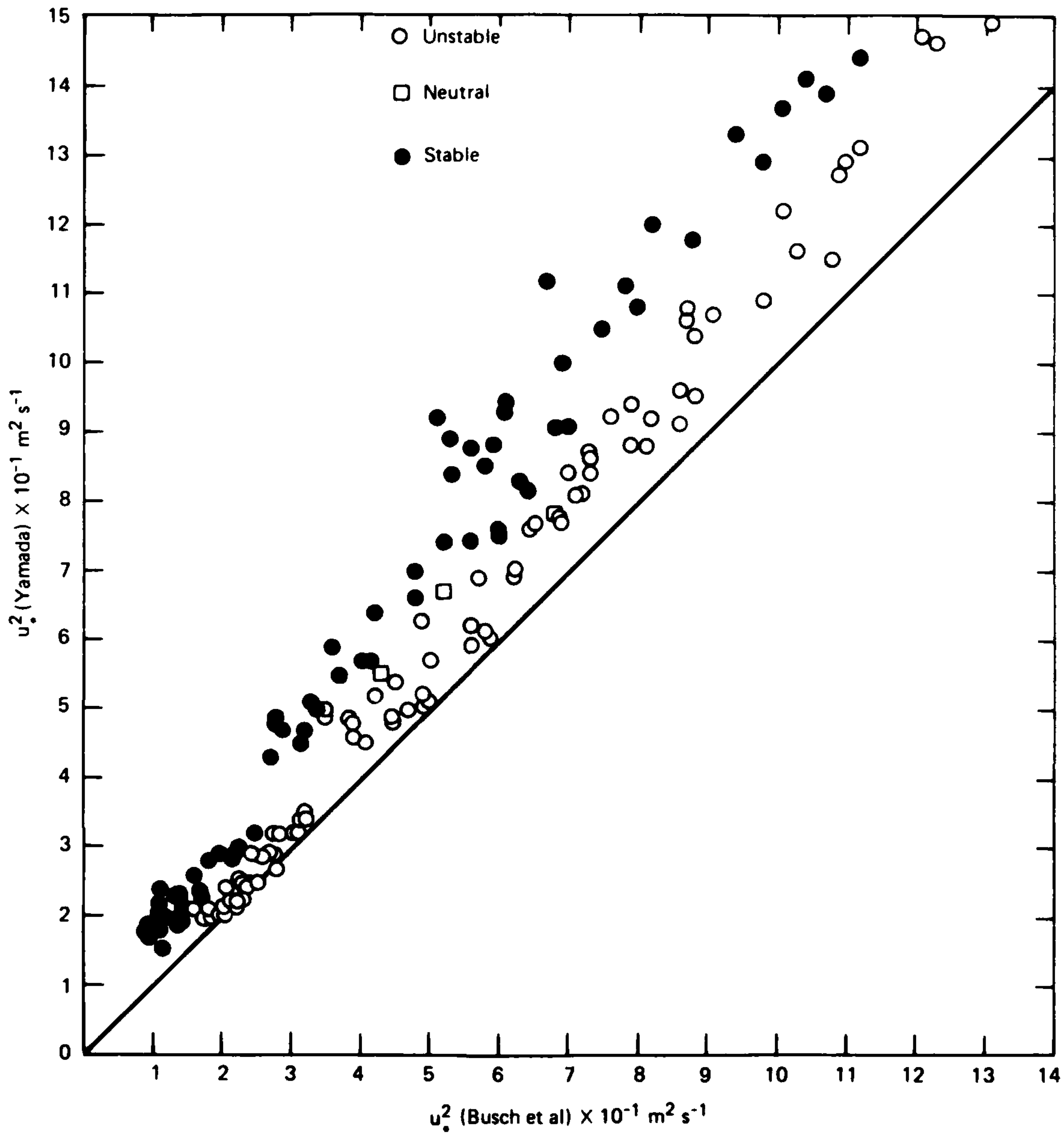

Fig. 1 - Surface stresses ($m^2 s^{-1}$) computed by generalized similarity theory versus those computed by the 1D PBL model for stable (•), unstable (o), and neutral (m) conditions. Points should fall on the diagonal line if two models agree.

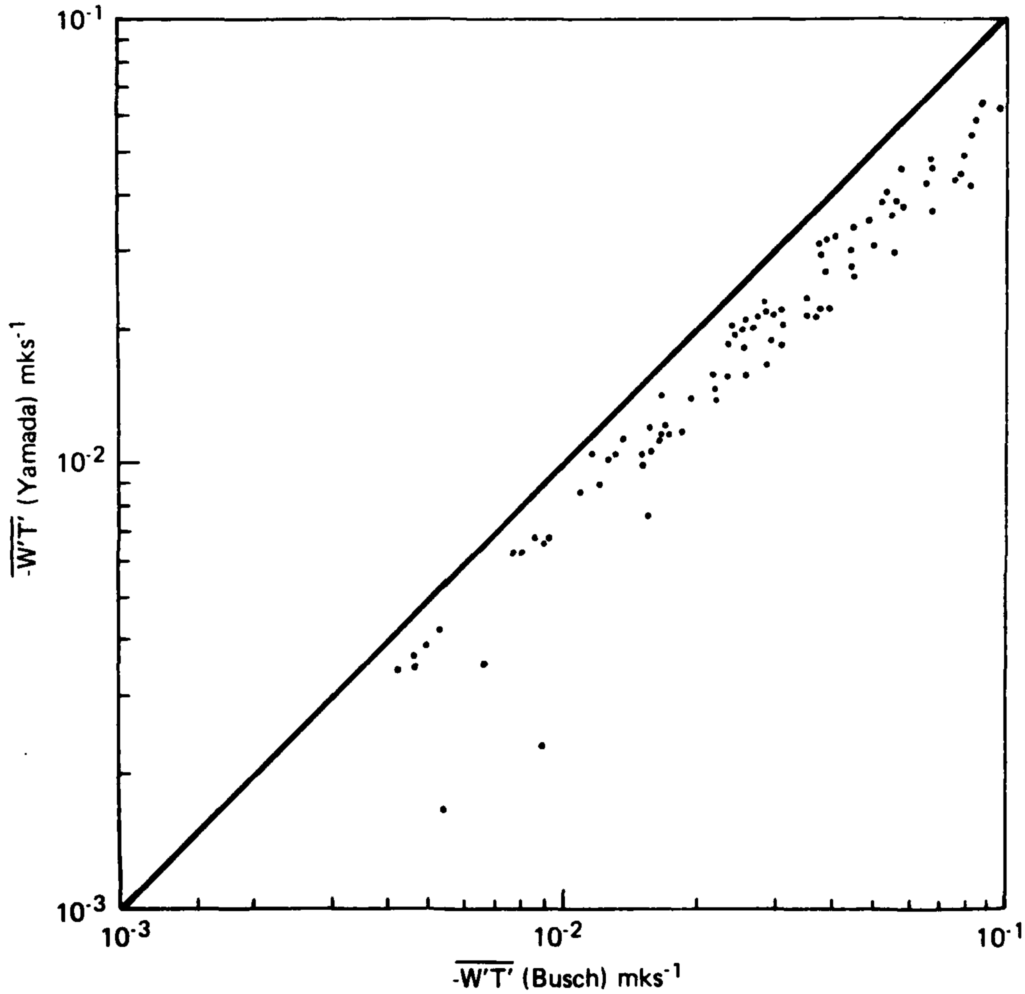

Fig. 2 - Same as Figure 1 except for upward surface heat flux in unstable condition (C m s^{-1})

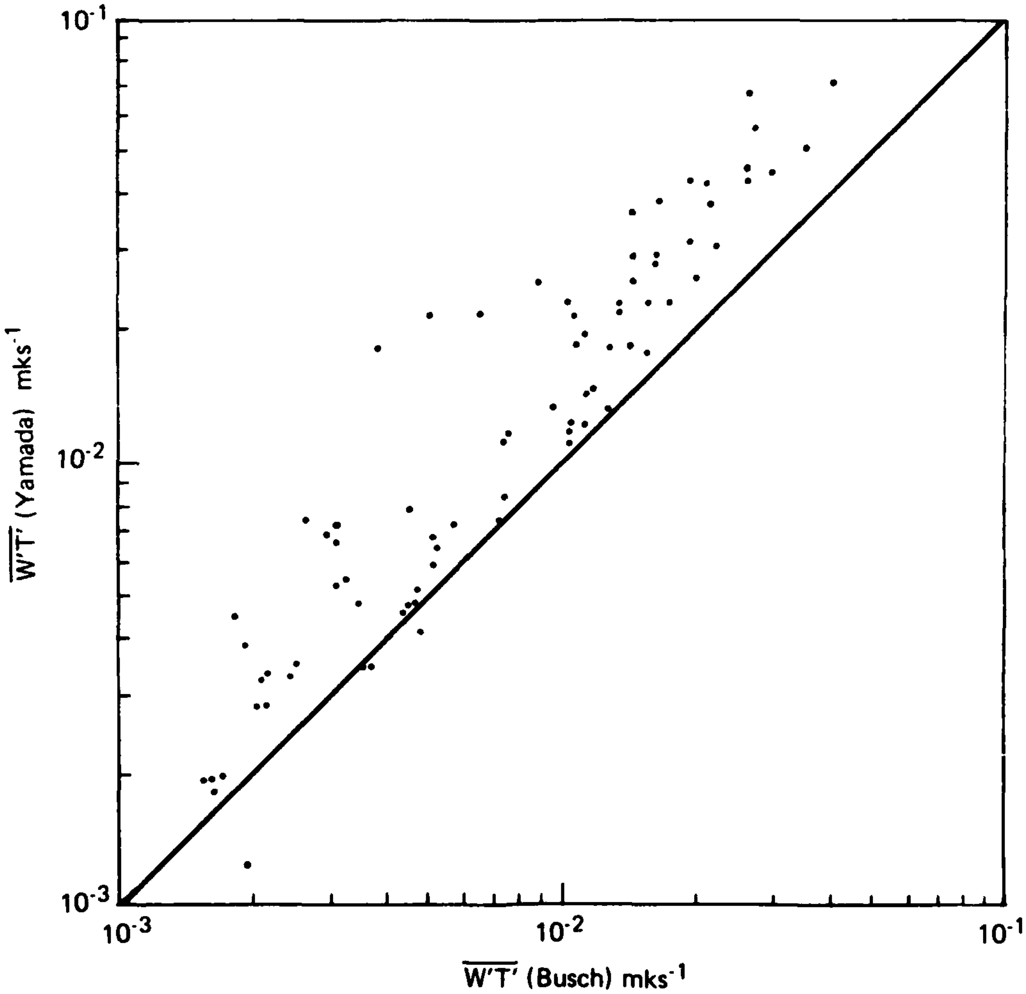

Fig. 3 - Same as Figure 1 except for downward surface heat flux in stable condition (C m s^{-1})

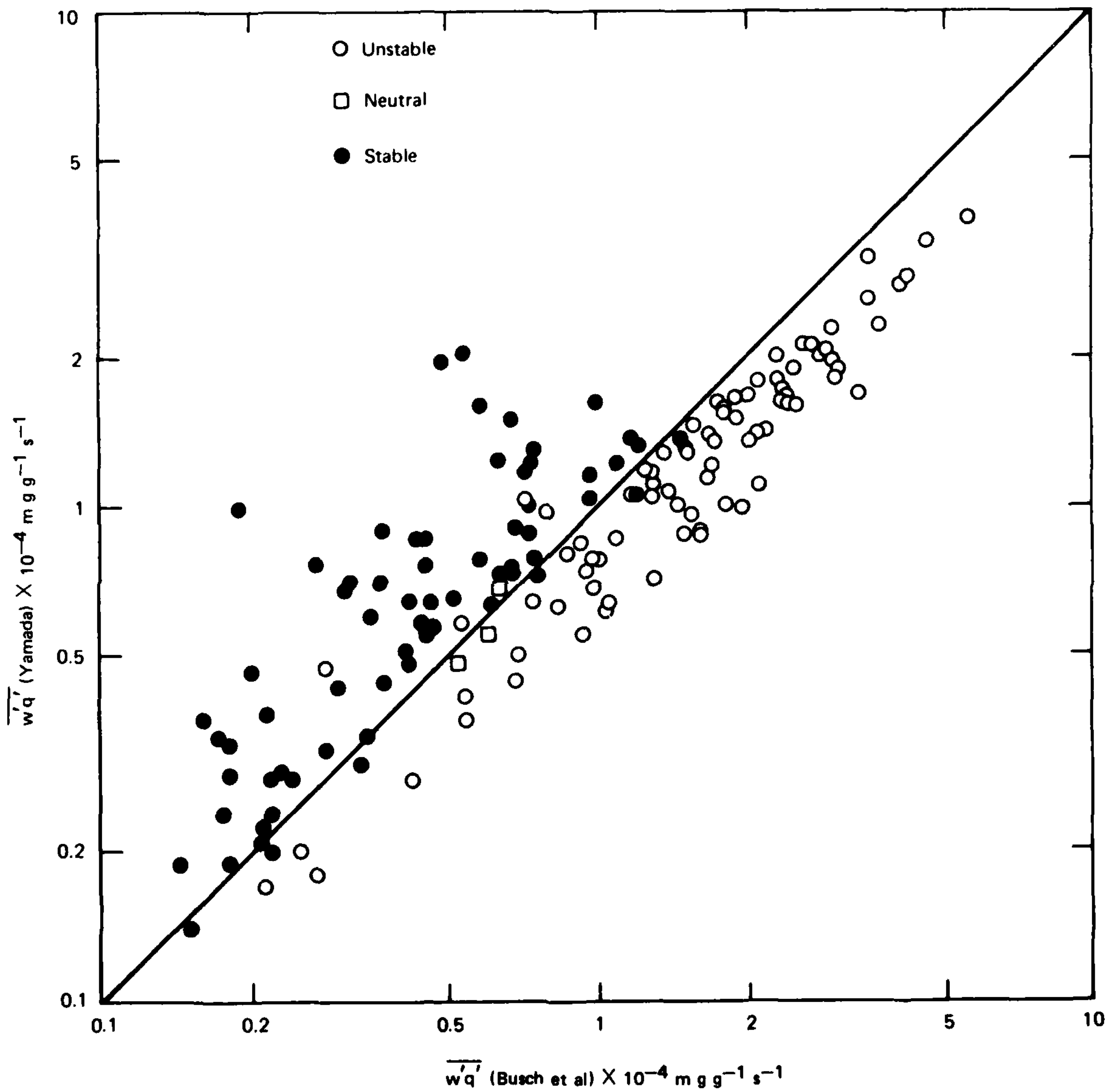

Fig. 4 - Same as Figure 1 except for surface vapor flux (g g^{-1} m s^{-1})

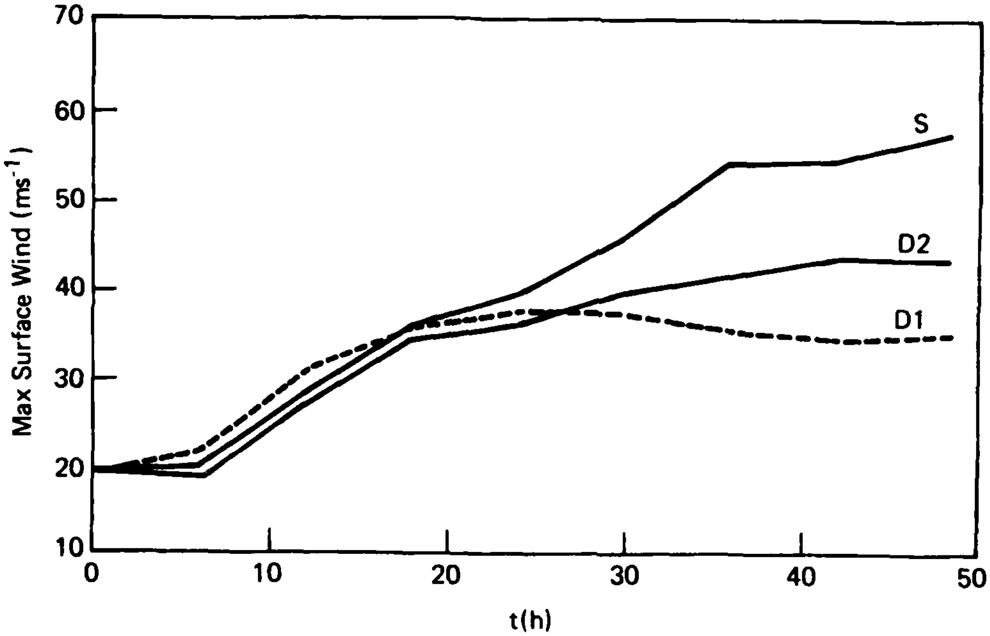

Fig. 5 - Time series of the maximum winds for model S, model D1 and model D2

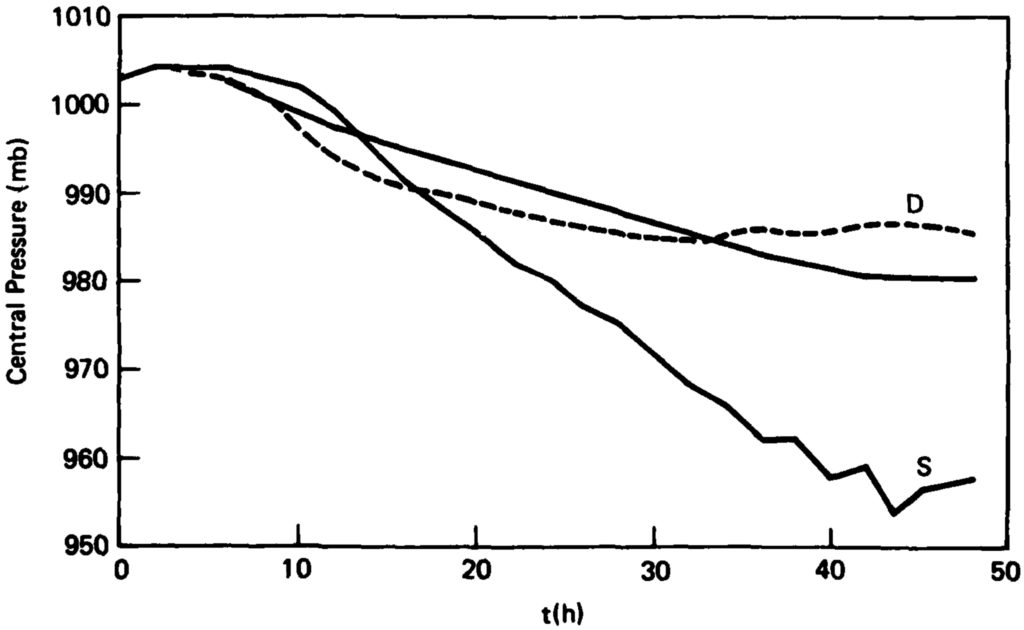

Fig. 6 - Time series of the minumum surface pressures for model S, model D1 and model D2

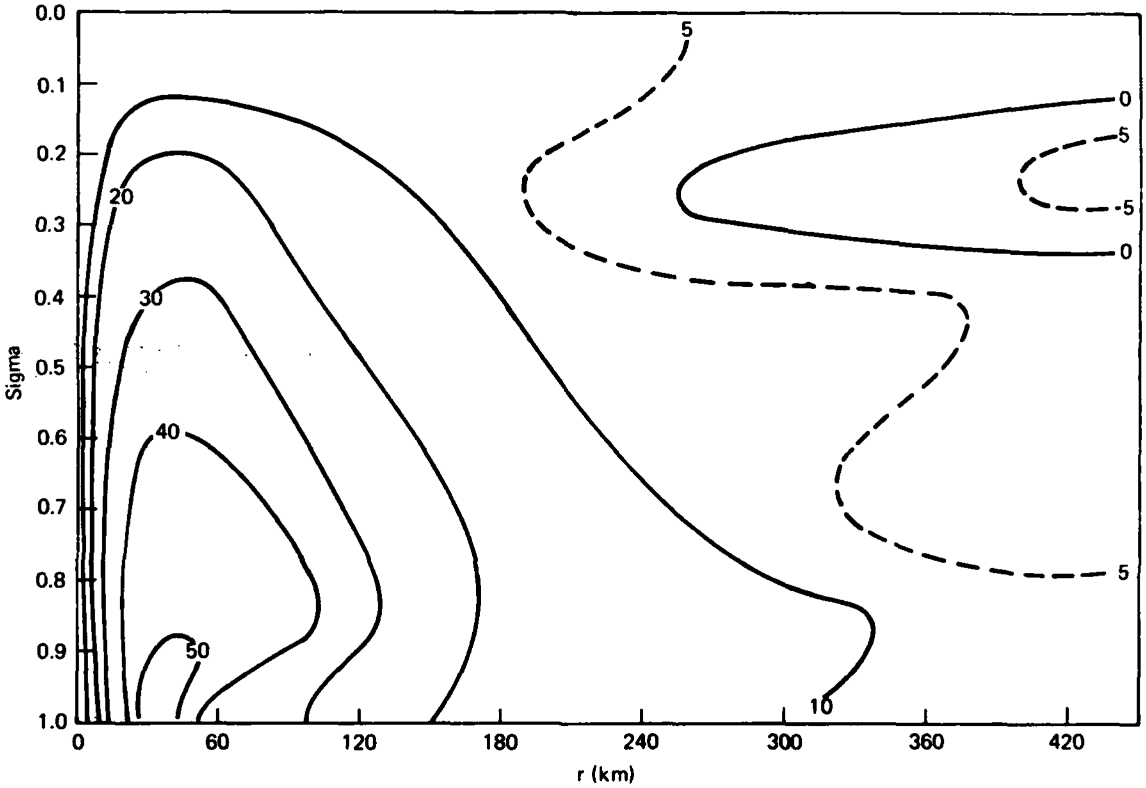

Fig. 7 - The quasi-steady tangential velocity (m s^{-1}) of model S at 48 h. Positive values denote cyclonic circulation, negative values denote anti-cyclonic circulation.

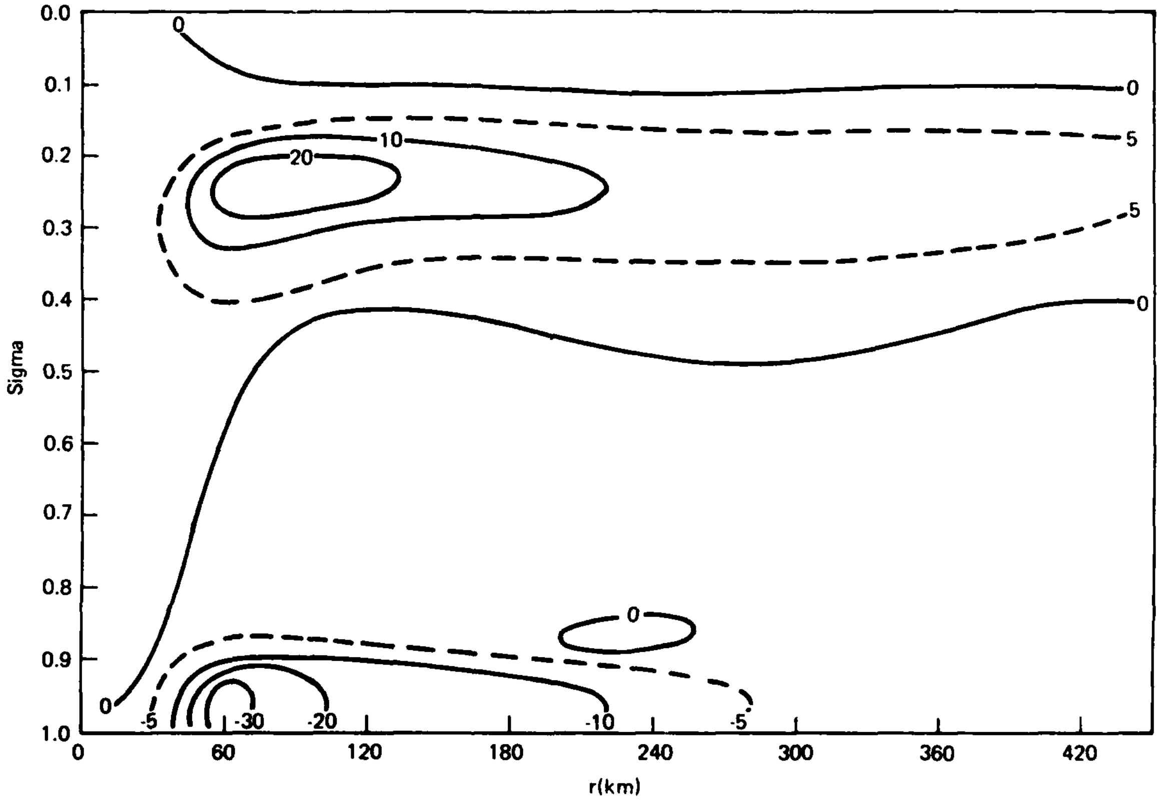

Fig. 8 - The quasi-steady radial velocity (m s^{-1}) of model S at 48 h. Positive values denote outflow, negative values denote inflow.

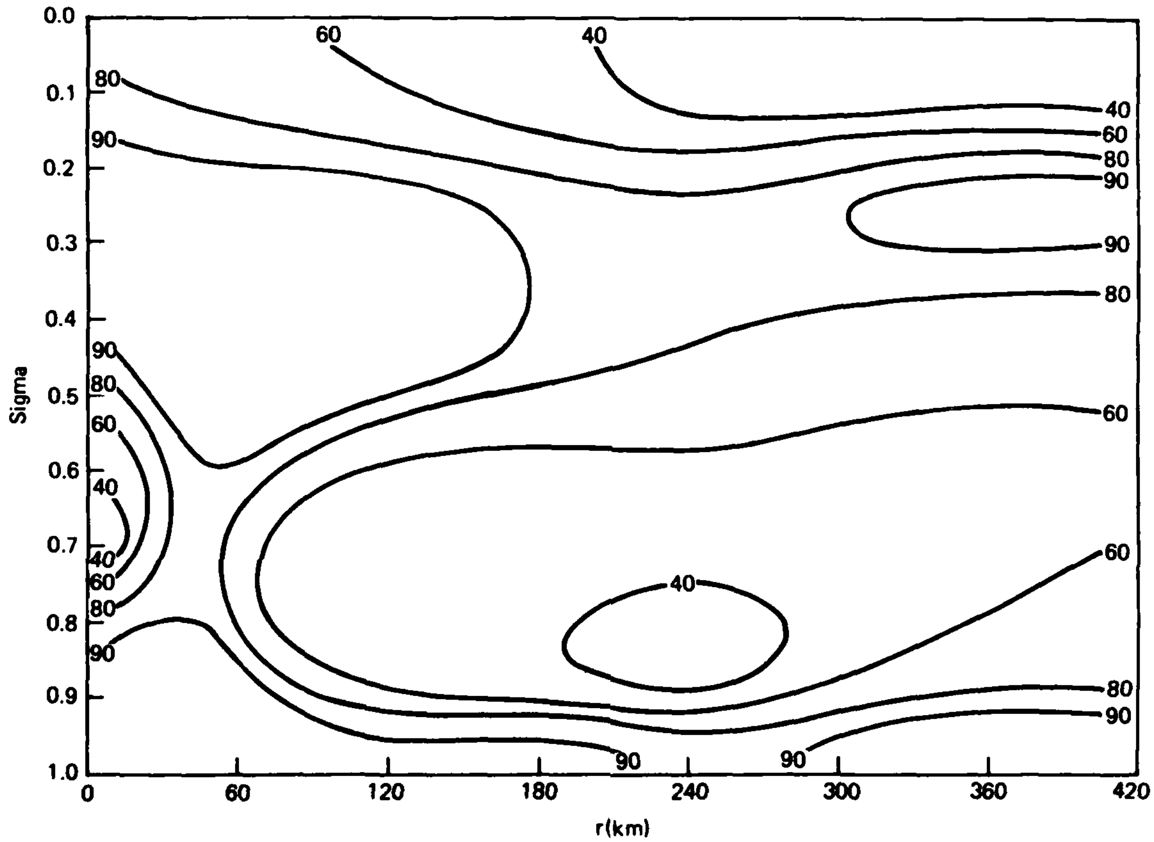

Fig. 9 - The quasi-steady relative humidity (%) of model S at 48 h

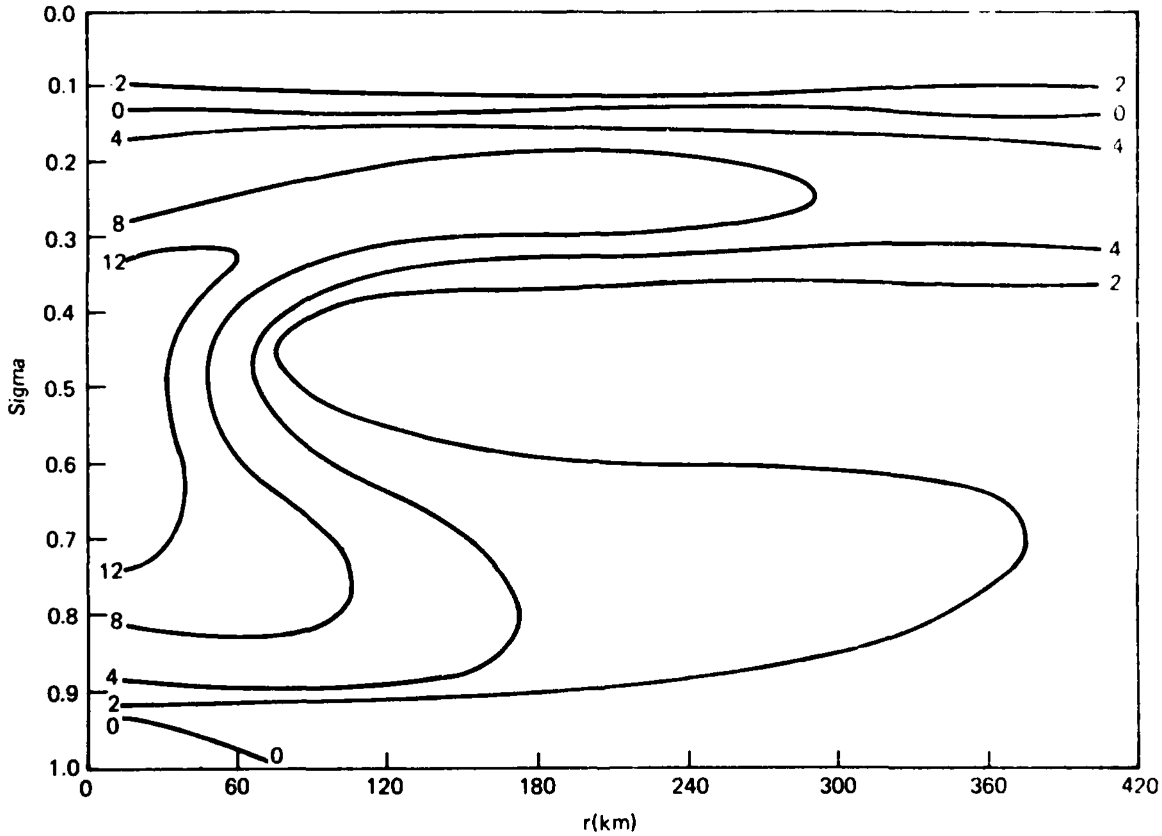

Fig. 10 - The quasi-steady temperature anomalies (C) of model S at 48 h. The anomalies are defined as temperature departures from the initial state.

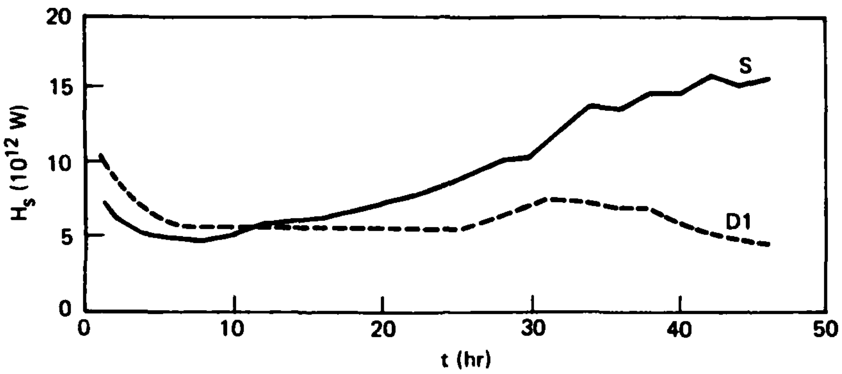

Fig. 11 - Time series of surface heat flux from r = 0 to 300 km of model S and model D1

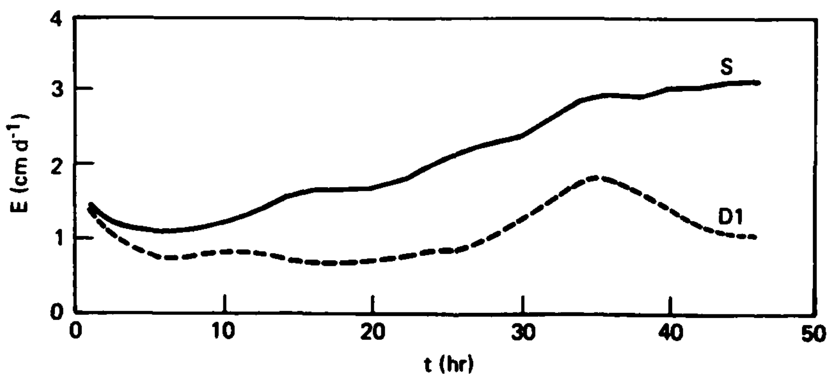

Fig. 12 - Same as Figure 11 except for averaged evaporation rate

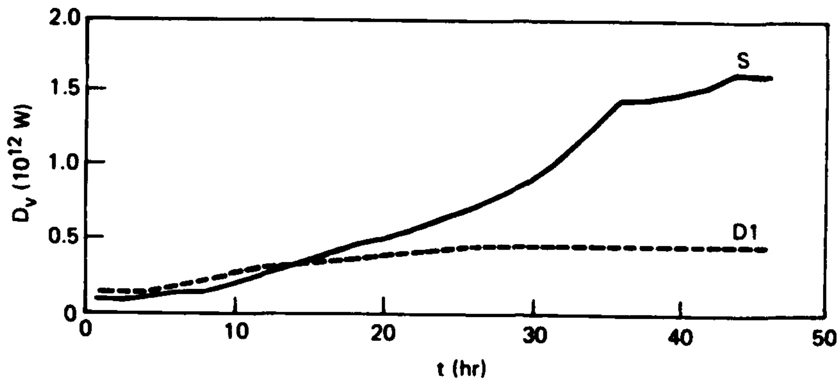

Fig. 13 - Same as Figure 11 except for KE conversion rate

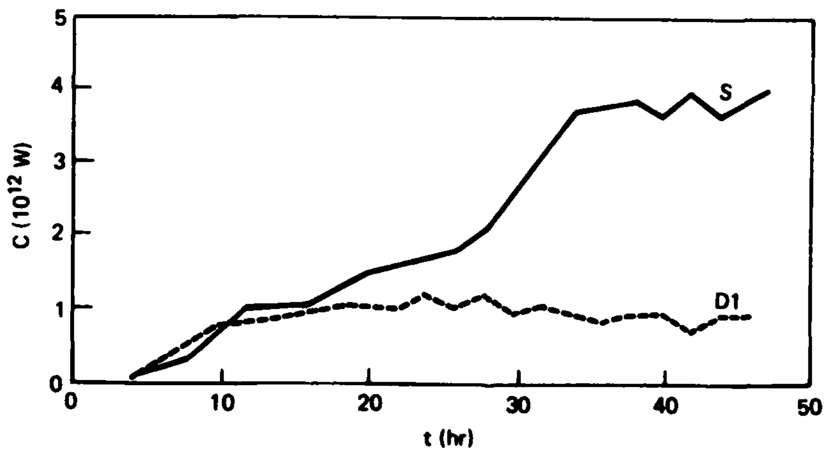

Fig. 14 - Same as Figure 11 except for KE dissipation rate due to surface friction

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